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Dynamic constraints on the crustal-scale rheology of the Zagros fold belt, Iran

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ABSTRACT

Thin-skinned fold-and-thrust belts are generally considered as the result of contractional deformation of a sedimentary succession over a weak decollement layer. The resulting surface expression frequently consists of anti- and synclines, spaced in a fairly regular manner. It is thus tempting to use this spacing along with other geological constraints to obtain insights in the dynamics and rheology of the crust on geological time scales. Here we use the Zagros Mountains of Iran as a case study as it is one of the most spectacular, well-studied thin-skinned fold-and-thrust belts in the world. Both analytical and numerical models are employed to study what controls fold-spacing and under which conditions folding dominates over thrusting. The models show that if only a single basal décollement layer is present underneath a brittle sedimentary cover, deformation is dominated by thrusting which is inconsistent with the data of Zagros Fold Belt. If we instead take into account additional décollement layers that have been documented in

the field, a switch in deformation mode occurs and crustal-scale folding is obtained with the correct spacing and timescales. We show that fold spacing can be used to constrain the friction angle of the crust, which is ~5 degrees in Zagros Fold Belt. This implies that on geological timescales, the upper crust is significantly weaker than previously thought, possibly due to the effect of fluids.

INTRODUCTION

It is often assumed that fold belts can be explained by folding of a sedimentary layer above a basal detachment formed by a weak layer. As the spacing between folds in such belts is quite regular, we can consider them as a large-scale natural experiment of crustal deformation. Ideally, it should be possible to combine fold spacing with other geological data and theory to constrain parameters such as crustal rheology that are difficult or impossible to constrain from field observations alone.

The classical explanation of folds in fold belts is that they are due to a folding instability, which is well known for a homogeneous sedimentary sequence with either a power-law viscous or an elastic rheology (Schmalholz et al., 2002; Burg et al., 2004; Schmalholz, 2006). The dominant wavelength λ_{dom}, for a viscous power-law layer of viscosity η_{sed} and with exponent n overlying a linear viscous layer of viscosity η_{salt}, is given by (Schmalholz et al., 2002):

$$\lambda_{dom} = 3.1\left(\frac{\eta_{sed}}{n\eta_{salt}}\right)^{\frac{1}{6}}\sqrt{\frac{H_{salt}}{H_{sed}}}H_{sed} \quad (1)$$

where H_{sed} and H_{salt} are the thicknesses of the sedimentary cover and of the salt, respectively. The growth rate (q_{dom}) of this instability non-dimensionalized over the background strain rate $\dot{\varepsilon}$ is given by (Schmalholz et al., 2002):

$$\frac{q_{dom}}{\dot{\varepsilon}} = 2.5n\left(\frac{\eta}{n\eta_{salt}}\right)^{\frac{1}{3}}\frac{H_{salt}}{H_{sed}} = 2.5n\left(\frac{\lambda_{dom}}{3.1H_{sed}}\right)^{2} \qquad (2)$$

and a combination of numerical and analytical studies have shown that $q_{dom}/\dot{\varepsilon}$ should be larger than ~20 for folding to form observable folds, rather than homogeneous thickening (e.g., Kaus et. al, 2008).

The Zagros Fold Belt of Iran constitutes a classical example of such a folded belt that is geologically (e.g., Stocklin, 1968; Alavi, 2004; McQuarrie, 2004; Sherkati and Letouzey, 2004; Mouthereau et al., 2007) and geophysically (e.g., Jahani et al., 2009; Hatzfeld and Molnar, 2010; Nissen et al., 2010) well studied due to excellent exposure and extensive seismic and borehole data from exploration. The main tectonic and stratigraphic units are summarized on Figure 1 and show that a particular feature of the Zagros Fold Belt is a consistent spacing of folds with a wavelength (λ_{dom}) of 14 ± 3 km. These folds are generally explained as detachment folding of the post-Cambrian sedimentary sequence above a basal weak layer constituted by the Hormuz salt.

The centroid depths of waveform-modeled earthquakes indicate that faulting is restricted to two structural levels located in the competent sediment cover units at 5–6 km depth and within the Precambrian basement at depth larger than 11 km down to depths of 30 km (e.g., Talebian and Jackson, 2004; Nissen et al., 2010). Seismic reflection profiles (Jahani et al., 2009) and field observations in the Fars region (Mouthereau et al., 2007) show a lack of major thrust faults cutting the folded cover up to the surface. This confirms that detachment folding rather than thrusting is the dominant deformation mode in the Zagros Fold Belt. In this aspect, the Zagros Mountains differ from other fold-and-thrust belts such as the Jura Mountains, where large-offset faults are continuous across the stratigraphic sequence, well imaged through seismic studies (Simpson, 2009).

Detachment folding theory should thus be perfectly applicable to the Zagros Fold Belt. Equations (1) and (2) show that fold spacing depends strongly on the rheology of the overburden and on the thickness of the basal salt layer. In the Zagros, a linear viscous overburden ($n = 1$) and a viscosity contrast of 100 between salt and overburden, requires a salt layer thickness of ~7.8 km to fit the observed spacing of folds (Equation 1). Yet, seismic data indicates that the thickness of the Hormuz salt is no more than 1 or 2 km (Jackson et al., 1990; Mouthereau et al., 2006; Jahani et al., 2009). If the sedimentary cover has a power law rheology instead, its power law exponent should be n~23 (Equation 2) to explain the data, which is considerably larger than estimates from rock creep experiments (Ranalli, 1995). Large power law exponents are often taken as evidence for a brittle rheology. Currently, however, there is no theory that can reliably predict the spacing of detachment folds in the case of a brittle overburden.

There is thus presently no satisfactory explanation for (1) why deformation in the Zagros Fold Belt is dominated by folding and not by thrusting and for (2) what controls the spacing of folds and how it is linked to crustal rheology. In order to address this, we performed thermo-mechanical numerical simulations to study the dynamics of detachment folding in the presence of a brittle sedimentary cover.

NUMERICAL MODEL

To study the effect of using visco-elasto-plastic rheologies on crustal dynamics, we have performed a series of numerical experiments using the finite element code MILAMIN_VEP (e.g., Kaus, 2010 and GSA Data Repository DR1). The viscosities of the weak layers are assumed to be linear and constant, which is a reasonable approximation for the rheology of salt.

The brittle layers have a temperature-dependent rheology of limestone (see DR1), which correspond to the majority of rocks within the sedimentary cover (Fig. 1C, Mouthereau et al., 2007). A linear geotherm of 25 °C.km^{-1} is initially applied (see DR1). For the low-temperature conditions of the Zagros Fold Belt, stresses are such that the rocks effectively deform in the brittle regime. Our model domain is initially 200x7.225 km in size (see DR1). The top boundary is a free surface with no erosion (see DR1) and a constant background strain rate of 10^{-15} s^{-1} is applied at the right of the model box, which results in 15% shortening after 5.5 Myrs consistent with geological constraints (see DR1). All other sides of the model have free slip conditions. Finally, to initiate the folding, the interface between the salt and the overburden rocks has random noise with maximum amplitude of 100 m. Model simulations are performed for 5.5 Myrs, after which results are interpreted.

RESULTS FROM NUMERICAL SIMULATIONS

With a 1.5 km-thick single basal detachment layer underlying a homogeneous brittle sedimentary cover, the models develop faults rather than folds (Fig. 2B). Such faults develop at early stages with a spacing that is approximately twice the brittle layer thickness. Subsequent deformation is locked around these folds that have a box-fold geometry. Compared to the Zagros Fold Belt, we thus obtain a too large wavelength and an incorrect deformation style. Additional simulations where we varied the frictional parameters of the crust, or the viscosity of the salt layer gave similar results (see Fig. DR1). We thus infer that it is impossible to reproduce the observed finite wavelength of Zagros Fold Belt folds (Fig. 1) by considering only one weak basal décollement layer, unless this layer has an unrealistically large thickness.

A detailed look at the stratigraphic column, however, reveals that the sedimentary cover is not rheologically homogeneous. Instead there are several layers that are composed of relatively weak rocks such as evaporites or shales (Fig. 1B, C, see detailed descriptions in McQuarrie, 2004; Sherkati et al., 2006 and Mouthereau et al., 2007). A second set of simulations took this fine-scale rheological structure into account (Fig. 3). The results are remarkably different from the previous experiments: rather than being fault-dominated, deformation is now achieved by folding, with a final wavelength similar to the one observed in the Zagros Fold Belt (Fig. 3). An analysis of the simulation shows that the spacing of the folds is fixed at a very early stage, after which the individual structures grow without clear geometric pattern, in accordance with field constraints (Mouthereau et al., 2007). The initial fine-scale rheological stratification of the sediment cover of the Zagros Fold Belt thus has a first-order effect on the development of upper crustal-scale structures. These results are in full agreement with a recent study of active seismicity in the Zagros Fold Belt which showed that both the Hormuz salt layer and the intermediate layers are mechanically-weak zones that form barriers to rupture for active faults (Nissen et al., 2010).

CONSTRAINTS ON CRUSTAL RHEOLOGY

The simulations presented in this study highlight the different modes of deformation that might occur in fold-and-thrust belts. However, they give limited insights into the underlying physics, as it remains unclear how sensitive the spacing of structures is to the rheology of the crust. For this reason, we developed a semi-analytical methodology drastically reducing the computational requirements that allows us to predict the outcome of numerical simulations in a large parameter space (see details in DR2). The resulting wavelength versus growth rate diagrams have a single maximum as a function of non-dimensional wavelength (Fig. 4A). Rigorously, these semi-analytical results are only valid for very small deformations. Yet, a comparison with numerical simulations reveals that they correctly predict the spacing of folds even after 5.5 Myrs, which confirms that fold-spacing is selected at a very early stage in the evolution of a fold and thrust belt (Fig. 4A).

Results for a homogeneous and brittle sedimentary cover reveal that the dominant growth rate is smaller than 20, which essentially means that folding will not be able to overcome background pure-shear thickening. Indeed, our numerical simulations indicate that this leads to fault-dominated deformation rather than folding (box folds, Fig. 2). If, on the other hand, weak layers are taken into account in the sedimentary sequence, the growth rate is significantly larger and the dominant wavelength is reduced (Fig. 4B). The addition of a single weak layer is sufficient to switch deformation styles from fault- to fold- dominated, and elasticity has a minor effect only.

Using the same semi-analytical methodology, we performed a large number of simulations and found that the two most important parameters are the viscosity of the salt/weak layers and the friction angle of the crust, whereas rock density plays little to no role. Plots of dominant wavelength and growth rate versus those two parameters show an approximate equal dependence on the two parameters (Fig. 4). The results also show that weak layers in all cases yields growth rates that are sufficiently large for the folding instability to dominate faulting.

In the case of Zagros Fold Belt, the effective viscosity of salt has been determined to be close to 10^{18} Pa.s, a value consistent with scaled laboratory-derived values (Spiers et al., 1990) and other modeling studies (Van Keken et al., 1993; Mouthereau et al., 2006). If we combine this with our method, we estimate that the effective friction angle for the crust in the Zagros Fold Belt on geological timescales is around 5°+/– 5° (Fig. 4B).

DISCUSSIONS AND CONCLUSIONS

Contrary to the common view of fold belts that often consider a single major basal décollement only, we demonstrate through the example of the Zagros Mountains that the whole stratigraphic sequence might influence the dynamics of the belt. Heterogeneities within the sedimentary cover, and weak layers in particular, control whether deformation is dominated by crustal-scale folds or by thrusts. The stratigraphy of a fold belt plays a much larger role than previously appreciated and should thus be taken into account if one wishes to reconcile field observations with physically consistent models of geological processes.

Balancing geological cross-sections in fold-thrust belts is a difficult exercise that aims at providing a consistent structural and kinematic interpretation of usually independent structural data. Our method paves the way for developing future generations of 2D and 3D dynamic reconstruction models for fold and thrust belts (e.g., Lechmann et al. 2010).

Moreover, we show that the regular spacing of folds puts constraints on the rheology of the crust on geological timescales. In the case of Zagros Fold Belt, the value for the friction angle we obtained in this manner is small (<10°), which indicates that the crust was rather weak, potentially due to large fluid pressures (e.g., Huismans et al. 2005).

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FIGURE CAPTIONS

Figure 1. Field constraints for the Zagros folded belt. A: Topography illustrates the regular spacing of folds with amplitude ~500–1000 m over on an area of ~80 000 km^2. Fold crest length

are of ~50 km in average. Inset shows the distribution of fold wavelengths measured for 88 anticlines from the Zagros Folded Belt. B: Cross-section (aa') across the Zagros Fold Belt based on field measurement (Mouthereau et al., 2007). λ corresponds to the average wavelength of the folds. This value is slightly smaller than the 15.8 +/− 5.3 km from Mouthereau et al. (2007) that took into account the folds from the whole Fars area. MFF and SF correspond to the seismogenic Mountain Front Fault and the Surmeh Fault, respectively, associated with basement faulting. Vertical fold velocity is 0.3–0.6 $mm.yr^{-1}$. C: Synthetic stratigraphic log where WL1, WL2 and WL3 correspond to the weak layers in the sedimentary sequence (Fm: Formation).

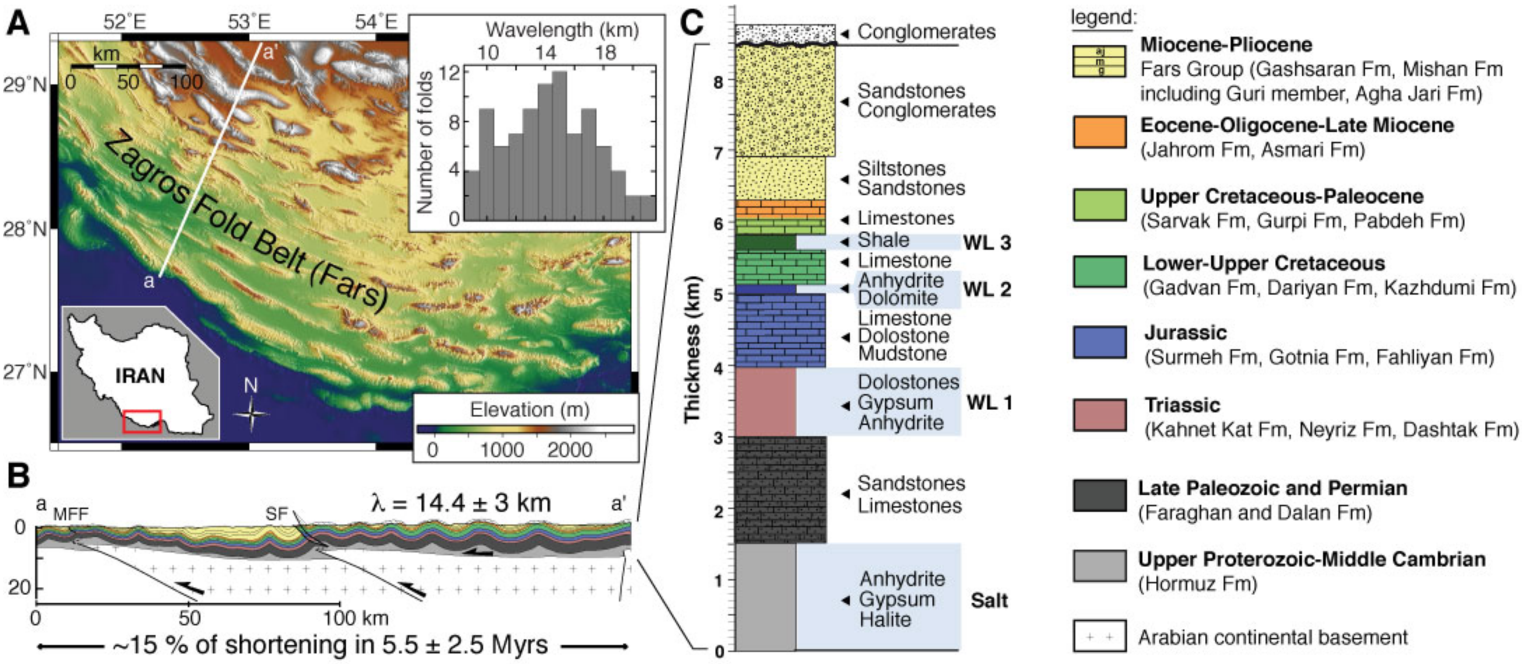
Figure 2. Simulation with a basal décollement layer only. A: Initial setup with a sedimentary thickness of 7.225 km. All rocks above the basal salt layer are homogeneous and have a friction angle of 5° and a cohesion of 20 MPa. A background strain rate of 10^{-15} s^{-1} is imposed at the right model boundary. B: Geometry, strain rate, and vertical velocities after 1.5 Myrs and 5.5 Myrs respectively. Deformation is localized along crustal-scale plastic shear zones and deformation structures are fault-dominated.

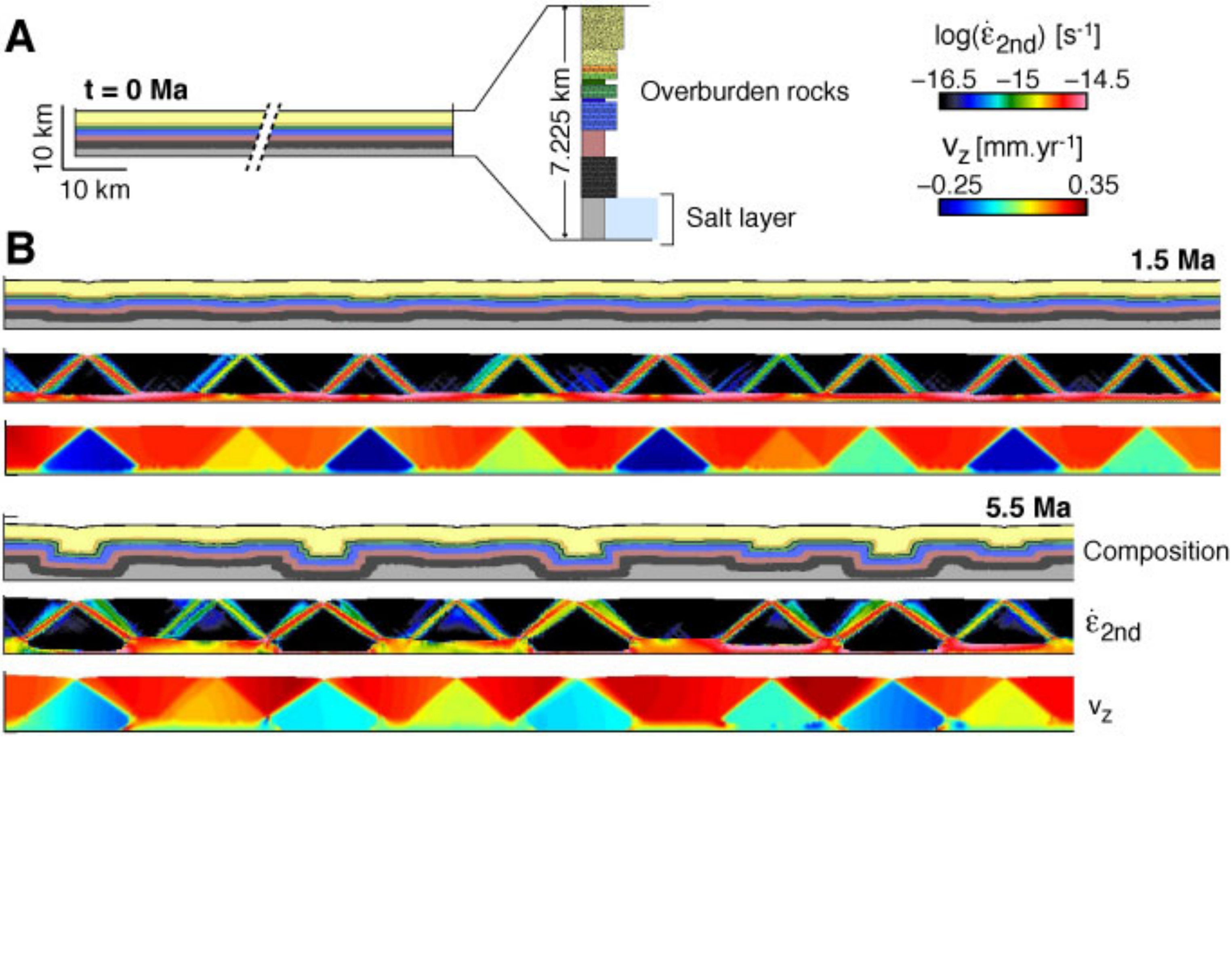
Figure 3. Simulation with intermediate crustal detachment layers. A: Initial setup as in figure 2, but with three additional weak crustal detachment layers with 10^{18} Pa.s. B: Snapshots of geometry, strain rate and vertical velocities at different times, which illustrate that crustal-scale folds rather than faults dominate the deformation pattern. Note that folds do not grow continuously with time, but rather grow to certain amplitude after which activity switches to a different fold.

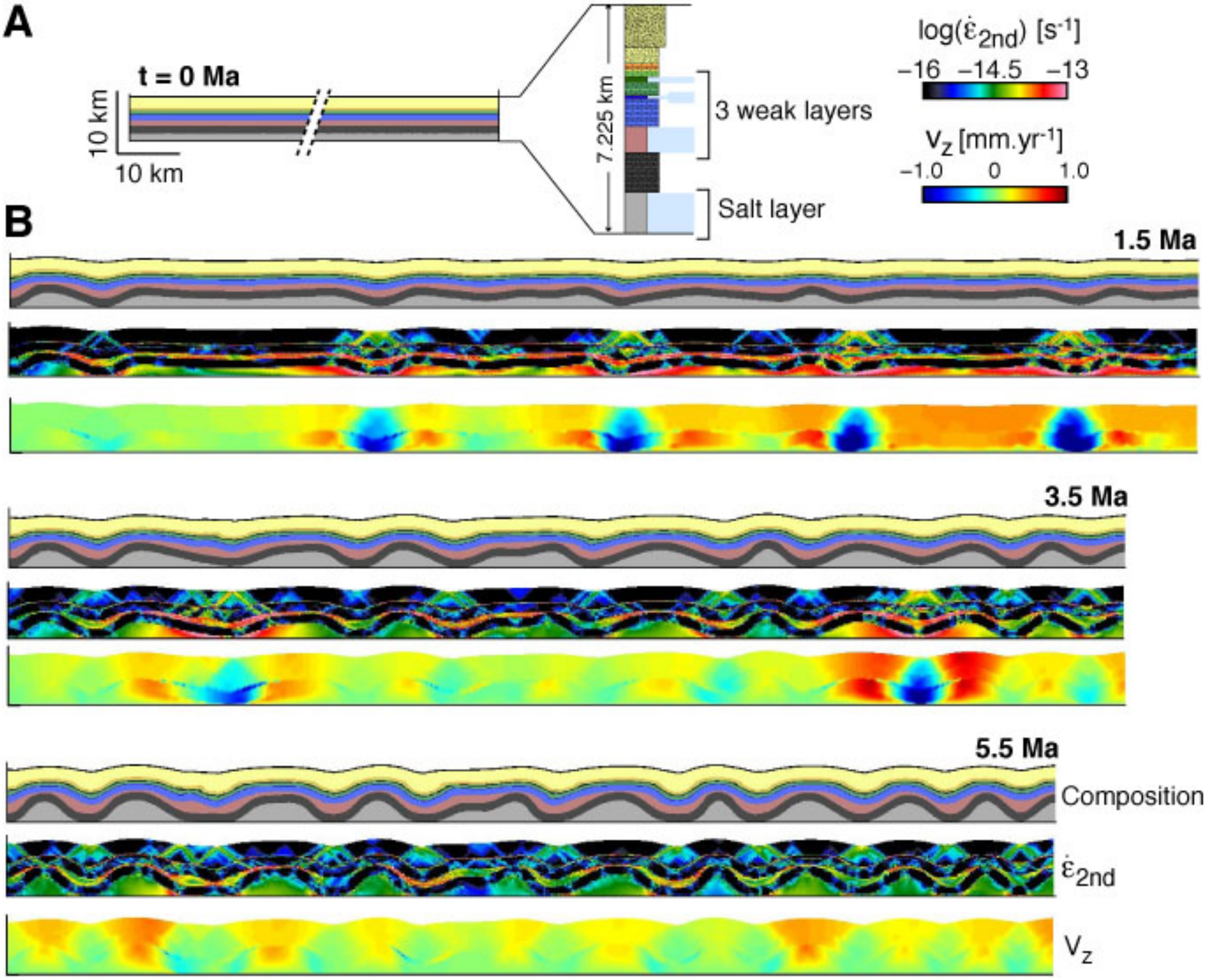
Figure 4. Influence of multiple weak layers and elasticity on folding. λ, H, q and $\dot{\varepsilon}$ corresponds to the wavelength, the total thickness, the growth rate and the background strain rate, respectively. VP and VEP correspond to visco-plastic and visco-elasto-plastic simulations,

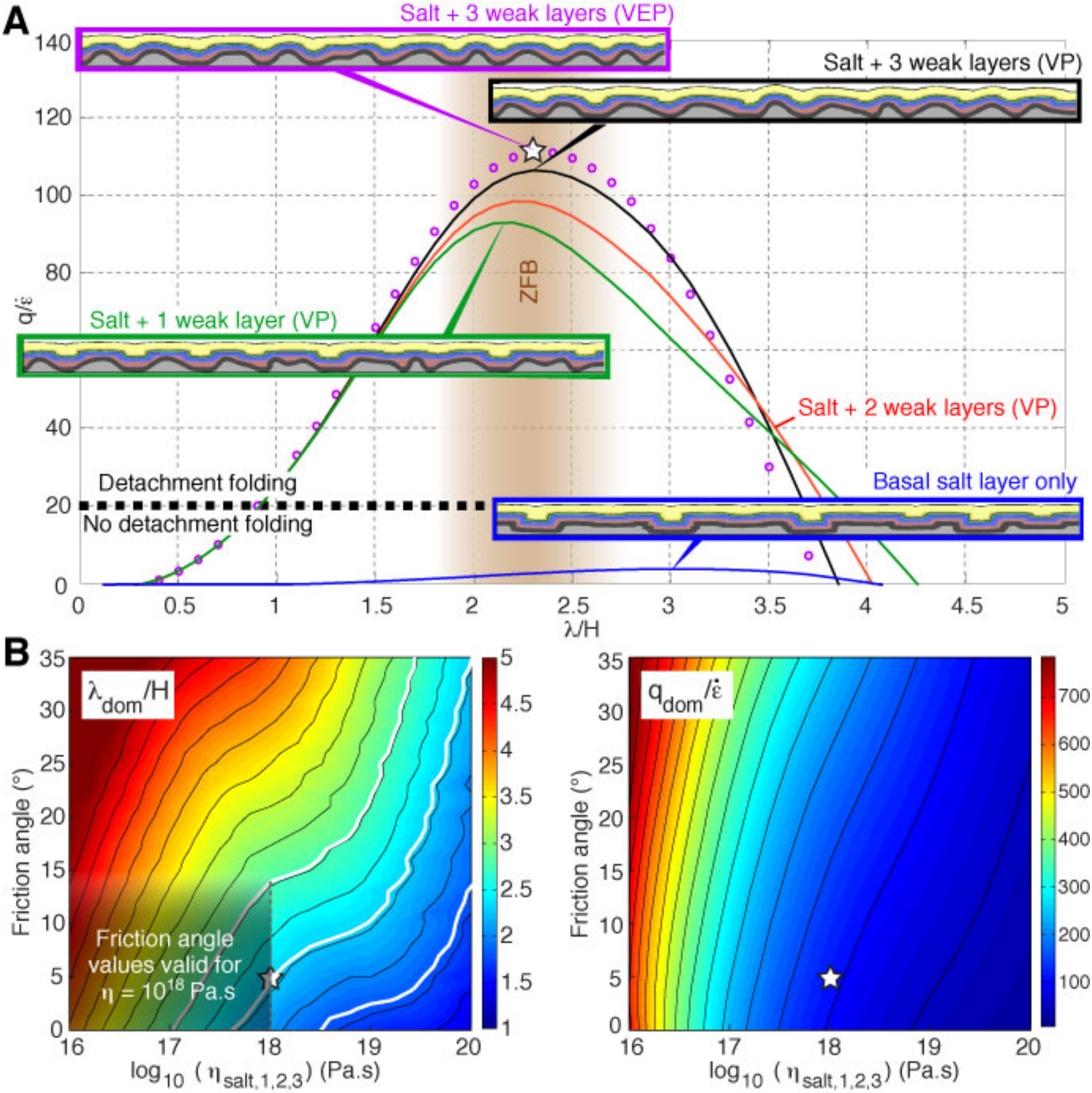
respectively. This diagram was produced using the semi-analytical approach described in DR2. A: Growth rate values obtained for given values of λ/H for 0, 1, 2 and 3 weak layers. For each case, the characteristic wavelength value corresponds to the highest value of growth rate (e.g., white star for the case with 3 weak layers). Insets show results of numerical simulations after 5.5 Myrs, which develop folds with a spacing that is in excellent agreement with the predicted characteristic wavelength. A single basalt décollement layer results in small folding growth rates and in thrust-dominated deformation. Addition of one or more weak layers to the brittle sedimentary cover results increases the growth rate significantly and leads to folding-dominated deformation. The ZFB brown area corresponds to the λ/H ratio of the Zagros Fold Belt. B: diagrams of characteristic wavelength (left) and corresponding growth rate (right) versus viscosity of the weak layers and friction angle of the crust. Thick white lines show the constraints for Zagros Fold Belt (average +/– 1 standard deviation). As in the Zagros Fold Belt salt viscosity is constrained independently, the best-fit friction angle for the crust is 5+/–5°.The white star corresponds to the simulation of figure 3.

[1]GSA Data Repository item 2011xxx, xxxxxxxx, is available online at www.geosociety.org/pubs/ft2011.htm, or on request from editing@geosociety.org or Documents Secretary, GSA, P.O. Box 9140, Boulder, CO 80301, USA.


A
52°E
53°E
54°E
29°N
28°N
27°N
km
0
50
100
a'
a
Zagros Fold Belt (Fars)
IRAN
N
Wavelength (km)
10
14
18
Number of folds
12
8
4
0
Elevation (m)
0
1000
2000
B
a
MFF
SF
λ = 14.4 ± 3 km
a'
0
20
0
50
100 km
~15 % of shortening in 5.5 ± 2.5 Myrs
C
Thickness (km)
Conglomerates
Sandstones
Conglomerates
Siltstones
Sandstones
Limestones
Shale
WL 3
Limestone
Anhydrite
Dolomite
WL 2
Limestone
Dolostone
Mudstone
Dolostones
Gypsum
Anhydrite
WL 1
Sandstones
Limestones
Anhydrite
Gypsum
Halite
Salt
legend:
Miocene-Pliocene
Fars Group (Gashsaran Fm, Mishan Fm including Guri member, Agha Jari Fm)
Eocene-Oligocene-Late Miocene
(Jahrom Fm, Asmari Fm)
Upper Cretaceous-Paleocene
(Sarvak Fm, Gurpi Fm, Pabdeh Fm)
Lower-Upper Cretaceous
(Gadvan Fm, Dariyan Fm, Kazhdumi Fm)
Jurassic
(Surmeh Fm, Gotnia Fm, Fahliyan Fm)
Triassic
(Kahnet Kat Fm, Neyriz Fm, Dashtak Fm)
Late Paleozoic and Permian
(Faraghan and Dalan Fm)
Upper Proterozoic-Middle Cambrian
(Hormuz Fm)
Arabian continental basement


A
t = 0 Ma
10 km
10 km
7.225 km
Overburden rocks
Salt layer
log($\dot{\varepsilon}_{2nd}$) [s^{-1}]
−16.5 −15 −14.5
v_z [mm.yr^{-1}]
−0.25 0.35
B
1.5 Ma
5.5 Ma
Composition
$\dot{\varepsilon}_{2nd}$
v_z


A
t = 0 Ma
10 km
10 km
7.225 km
3 weak layers
Salt layer
log($\dot{\varepsilon}_{2nd}$) [s^{-1}]
−16
−14.5
−13
V_z [mm.yr^{-1}]
−1.0
0
1.0
B
1.5 Ma
3.5 Ma
5.5 Ma
Composition
$\dot{\varepsilon}_{2nd}$
V_z


A
Salt + 3 weak layers (VEP)
Salt + 3 weak layers (VP)
Salt + 1 weak layer (VP)
Salt + 2 weak layers (VP)
Basal salt layer only
ZFB
Detachment folding
No detachment folding
q/ε̇
λ/H
B
λdom/H
qdom/ε̇
Friction angle (°)
Friction angle values valid for η = 10^18 Pa.s
log10 (ηsalt,1,2,3) (Pa.s)